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THE EFFECTIVE AEROSOL ATTENUATION ON DETERMINATION ERROR
OCEAN AND ATMOSPHERE TEMPERATURES BY REMOTE METHODS

Academy of Sciences USSR

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ANNOTATION

We will discuss data existing in literature on errors in establishing the temperature of the underlying surface and vertical profile of temperature according to deteriorating radiation of the Earth, in the $15\text{ }\mu\text{m}$ CO_2 absorption band and in the $10\text{-}12\text{ }\mu\text{m}$ "transparency window" which involves the fact that aerosol attenuation was not considered. The experimental data reveal the significant contribution of the aerosol to total radiation attenuation whereas, according to the results of model calculations, the errors in establishing temperature caused by the aerosol do not exceed the infrared in most cases. This disparity is explained by the existence of fairly dense aerosol formations (transitory in optically thin cloudiness) in the cold layers of the atmosphere. Evaluations are made of the relationship of magnitude of establishment error to location and optical thickness of the aerosol layer (numerical experiment),

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Introduction

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Experience in determining the temperature of the underlying surface T_0 and the vertical profile of the temperature of the atmosphere $T(\xi)$ according to measurement of intrinsic Earth radiation in the infrared range of the spectrum from satellites (see, for example, [1,2]), attests to the strong effect of aerosol attenuation of infrared radiation on precision in solving these problems (ξ is altitude in relative units of pressure). This same conclusion follows from an analysis of results of ground and spacecraft measurements of spectral transmission of solar radiation of the atmosphere in "windows" in the infrared range of the spectrum [3,4], and also from analysis of certain aircraft and spacecraft measurements of intrinsic radiation of the atmosphere in these windows [5, 6].

Errors in determining T_0 and $T(\xi)$ involving neglect of aerosol attenuation in the 10-12 μm window and in the 15 μm CO_2 band, according to this data, can exceed 10°C for a visually cloudless atmosphere. According to the data of this same work [7,8], in which aerosol atmosphere models are used, based on measurements of vertical profiles of concentrations of particles [9, 10], which indicate that errors do not exceed 1°C (even for a very murky atmosphere, determination error for T_0 can reach 2°C).

*Numbers in the margin indicate pagination in the foreign text.

Such significant differences in the estimates of the effect of aerosol attenuation in the problems of a hermetically sealed probe of the atmosphere require consideration of the possible causes for these differences. This is the purpose of this article. However, before getting into such an analysis, it is necessary to propose fairly specific criteria which would make it possible to delimit the concept of the aerosol and optically thin cloud layers existing in the atmosphere. As such a criterion, one can take the maximum difference in variation of optical thickness of a vertical column of the atmosphere $\Delta\tau_v$, usually obtained in ground measurements of spectral transmission of radiation of a cloudless atmosphere in the infrared range "windows" for each fixed value of atmosphere containing moisture W with fairly close vertical profiles of temperature $T(\xi)$ and moisture $q(\xi)$ (v is frequency). As the results of these measurements show [2,4], for the 10-12 μm window, $\Delta\tau_v$ reaches a value of 0.3-0.4 μm when $W = 1.4 \text{ cm}$. /2

Let us note that the minimum value of optical thickness of a vertical column of atmosphere which can be considered as the maximum possible (with the indicated values of W) values of optical thickness of water vapor τ_w consist of 0.1-0.4 for the 10 μm and 0.15-0.6 for the 12 μm . Thus, as the upper boundary of optical thickness of the aerosol $\tau = \Delta\tau_v$ in 10-15 μm spectrum one can assume for definiteness, $\tau_a \text{ max} = 0.4$.

Existing Estimates of Aerosol Effects

Estimates of the effects of aerosols on intrinsic radiation of the atmosphere in the 15 μm CO_2 band field, obtained in reference [7], were based on calculations of radiation intensity $I_{\Delta v}$ in seven spectral intervals of the 15 μm CO_2 band (668.7, 679.8, 692, 701, 709, 734, 750 cm^{-1}) and in the

899 cm^{-1} window, which were used in the satellite infrared radiometer, the Nimbus-III, for determining $T(\xi)$. Calculations were made for models of bright (molecular) and cloudy atmospheres. Measurements of vertical profiles of particle concentrations of aerosol $N(\xi)$, presented in [9] up to an altitude of ≈ 30 km for three stations at high, average and low latitudes, and intermediate measurements of the vertical profile of the coefficient of aerosol attenuation $\beta(\xi)$ at a wavelength of $\lambda = 0.55 \mu\text{m}$ in the 3-34 km layer of atmosphere [IC [iskusstvennyy sputnik, artificial satellite]] were taken as the basic model for a cloudy atmosphere. According to these measurements and according to radio sounding balloon data [7], four models of vertical profiles were constructed $N(\xi)$, $T(\xi)$ and $q(\xi)$; and here, in all the profile models $N(\xi)$ for the layer $\xi = 0.6-1$, by extrapolation N from $\xi = 0.6$ to $\xi = 1$ according to the exponential law to a given value $N(1) = 1840 \text{ cm}^{-3}$ was constructed corresponding to visibility distance $L = 2$ km when $\lambda = 0.55 \mu\text{m}$. The optical parameters (coefficients of attenuation, optical thickness and indicatrix of diffusion) for these models of the atmosphere were calculated using distribution of particles by dimension, characteristic for a "maritime" aerosol model [11], which gives the highest value of optical thickness of the aerosol for a given concentration of particles $N(\xi)$. /3

For a spectral range of 668-750 cm^{-1} , the optical thickness of aerosol absorption varied for the model [9] in a range of 0.12-0.14 and model [10] -- in limits of 0.17-0.18.

The optical thicknesses of aerosol diffusion in this range for both models amounted to a value on the order of 0.01. For the 899 cm^{-1} window, the value τ_a approximately equals 0.14 was taken. Calculation of $I_{\Delta v}$ in reference [7] for these atmospheric models showed that disregarding aerosol attenuation

leads to a decrease in $I_{\Delta v}$ of not more than 0.6% in the CO_2 band (757 cm^{-1}) and not more than 1.6% in 899 cm^{-1} window. Mean quadratic errors in determining $T(\xi)$ caused by such errors in $I_{\Delta v}$, amount to 0.4°C for the $N(\xi)$ profile [10] and 0.7°C for the $N(\xi)$ profile [9]. Maximum errors reach 1.7°C for $\frac{1}{4}$ the smallest mean quadratic error 2.5°C in determining $T(\xi)$ by a method of minimum information used [7].

Similar results were obtained in reference [8] for evaluating error in determining T_0 for radiation and the $11.2 \mu\text{m}$ window caused by aerosol attenuation. The authors of reference [8] made calculations of $I_{\Delta v}$ in this window for 135 models of a cloudy atmosphere constructed in the following way.

1. Distribution of particles by dimension were taken according to Young

$$\frac{dN}{dr} = cr^{-v} \quad (1)$$

for three values of the parameters and three ranges of measurements of particle radii r , for which, in reference [12], all of the optical parameters of the aerosol are calculated.

2. Vertical profiles of the concentration of aerosol particles are taken in reference [18] for models of a bright, cloudy and very cloudy atmosphere characterized by values of visibility at a distance from the surface of Earth $L = 23; 5$ and 1 km , respectively, and besides the last model is constructed by extrapolation of the value of N for a cloudy atmosphere with a level of 1 km at the value N from the Earth's surface, corresponding to $L = 1 \text{ km}$.

Optical thicknesses of the aerosol τ_a for these models in the $11.2 \mu\text{m}$ window reached values of $0.066, 0.216, 0.451$ for bright, cloudy and very cloudy atmospheres.

3. Each of the 27 models of the aerosol atmosphere are combined with five vertical profiles of temperature characterized by tropical, central and high latitudes for summer and winter; for this, in reference [8] $I_{\Delta V}$ was calculated at a level of 20 km.

The differences between radiation temperature T_r , corresponding to calculations of $I_{\Delta V}$, and initial temperature of the Earth's surface T_0 , the characteristic errors can be caused by the existence of aerosol attenuation of intrinsic radiation of Earth in the window being considered. According to analysis in [8] in the very unfavorable case of a very cloudy atmosphere when $V = 3.5$ and the range $r = 0.08-10 \mu m$, this error reaches $2^\circ C$, that is, close to the estimate in reference [7], although the maximum τ_a in these references differs by more than two times and the concentration of particles at the Earth's surface by a magnitude of 2. /5

It is not difficult to discover the cause for small errors in determining $T(\xi)$ and T_0 caused by aerosol effects and the low sensitivity of deteriorating radiation to such large variations in aerosol parameters. It is common in the character of vertical distribution of concentration of particles in models of cloudy atmosphere taken in references [7,8]. The point is that with such profiles of $N(\xi)$ practically all of the aerosol is concentrated in the boundary layer of the atmosphere with thickness 1 km ($\xi = 1-0.9$), and consequently the optical thickness $\tau_a(1, 0.9)$ of this layer is close to full optical thickness τ_a , and optical thickness of the overlying layer $\tau_a(0, 0.9) \approx 0$. Then, temperature $T(\xi)$ in the boundary layer in the models considered in [7,8] change comparatively little (not more than $5-7^\circ C$).

If, for these conditions, one considers the process of transformation of radiation of the Earth's surface $B_v(T_0)$ with its passage through an aerosol boundary layer of the atmosphere (maximum effect of such transformation will be, naturally, detected in "transparency windows"), then one can say that the aerosol attenuation $B_v(T_0)$ is almost completely compensated for by aerosol radiation of the boundary layer and this compensation does not depend on the degree of cloudiness of the latter.

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Indeed, in the expression

$$I_{av}(0) = B_{av}(T_0)e^{-\tau_a} - \int_0^{\xi} B_{av}[T(t)] \frac{\partial}{\partial t} (p_{av}(t) \exp[-\tau_a(0,t)]) dt \quad (2)$$

coupled with

$$I_{av} \approx B_{av}(T_0), T(\xi)$$

and optical thickness of the $\tau_a(0,t)$ layer from the current level t to the upper boundary of the atmosphere $\xi = 0$, the integral member which describes radiation of the atmosphere can be presented in the form of the second component:

$$\int_0^{\xi} B_{av}[T(t)] \frac{\partial}{\partial t} \exp[-\tau_a(0,t)] dt + \int_0^{\xi} B_{av}[T(t)] \frac{\partial}{\partial t} \exp[-\tau_a(0,t)] dt$$

(for simplicity, the "window" will be considered where $P \equiv I$).

For the models of an aerosol used in [7,8], the second component which describes radiation of the layer of atmosphere above the boundary, will be somewhat neglected in comparison with the first member which describes radiation of the boundary layer of the atmosphere. Radiation of this layer can be presented in the form

$$I_{av}(0,g) \approx B_{av}(\tilde{T}) \{-\exp(0;0,g) + \exp(-\tau_a)\} \quad (3)$$

where T is the temperature on a certain level of the boundary layer close to T_0 .

Then, with a certain degree of precision, from formulas (2) and (3) we find:

$$I_{\Delta v}(0) \approx B_{\Delta v}(\tilde{T}) \exp[-\tau_a(0;0.9)] \approx B_v(\tilde{T}) \approx B_v(T_0)$$

from which, obviously, follows the proximity obtained in [7,8] of radiation temperature T_r (1) to P_0 . From the relationships (2) and (3), it is obvious that the error $\Delta T = T_r - T_0$ will decrease with a decrease in the temperature gradient of the boundary layer with a fixed state of the atmospheric model aerosol or with a decrease in optical thickness of the layer of the atmosphere lying above the boundary layer with a fixed profile of $T(\xi)$. Such relationships for ΔT are just like 7 those obtained in [8]. It is completely obvious that with such a consideration of relationships (2) and (3) for the $15 \mu m$ CO_2 band, the aerosol effect in the atmospheric models [7,8] will be as small as before, inasmuch as in the fields of fairly high absorption of CO_2 molecules, radiation $I_{\Delta v}$ will be generated by high layers of the atmosphere with a low concentration of aerosol.

On the basis of the analysis presented and a comparison of the results of calculations [7,8] with materials for determining the temperature analysis according to measurements of $I_{\Delta v}$ in the $10-12 \mu m$ window which indicate [illegible] errors of $\Delta T = T_r - T_0$ exceeding $2^\circ C$ for a visually cloudless atmosphere (see, for example, [1,2]), can lead to the following conclusion. The most probable cause of large errors of thermal probing of the atmosphere can be the existence of aerosol layers with fairly large optical thickness at different altitudes in low temperature fields (for example, at

levels of condensation, in the upper troposphere, in the tropopause). Then, due to the large difference in temperature on the Earth's surface and at these levels, the aerosol effects can be significant.

Analysis of the Effects of Aerosol Layers

The layers of atmospheric particles formed are caused by existing temperature inversion at different levels and are characterized by stratification of the atmosphere under the inversion [13]. The effect of such layers on intrinsic radiation of the atmosphere was observed from aircraft with simultaneous measurements of vertical profiles of atmospheric radiation $I_{\Delta V}$ in the 10-12 μm window and the coefficient of diffusion σ in the visible field of the spectrum with maximum σ at 1-2 km levels [5]. Layers were also found at altitudes of 1-2 and 4-6 km when determining vertical profiles of optical thickness $\tau_a(\xi)$ according to measurements of angular distributions of $I_{\Delta V}(\theta)$ from onboard a spacecraft in the tropical region of the Atlantic [6]. /8

Such layers are the inevitable forerunners of optically thin clouds which are the largest source of uncontrolled errors in determining $T(\xi)$ and T_0 , inasmuch as such clouds cannot be detected according to measurements of radiation from satellites or even by pictures in the visible or infrared zones of the spectrum. For this reason, the criteria mentioned in the introduction of division of aerosol layers and optically thin clouds is conditional in the concept that the analyses presented below of error in determining $T(\xi)$ and T_0 , due to the aerosol, can be considered as estimates of similar effects of thin cloudiness. It is clear that the methods of calculation of aerosol effects in problems of thermal probing of the atmosphere must automatically provide a calculation of analogous effects of an optically thin cloudiness.

For estimates of error in establishing $T(\xi)$ involving noncalculation of the aerosol effect on deteriorating radiation, the following numerical experiment was conducted. According to a given profile of $T(\xi)$ and values of optical thickness of the aerosol layer τ_a , located at a given height, according to formula (2), the values of $I_{\Delta V}$ were calculated and in 6 intervals of $15 \mu\text{m}$ of the CO_2 bands with centers: 670, 690, 702, 738, 760, 830 cm^{-1} transmission functions for CO_2 $P_V^{\text{CO}_2}(\xi)$ calculated with averaging for 5 cm^{-1} taken from reference [15]. In all cases, it was considered that the underlying surface is absolutely black and its temperature coincides with the temperature of the lower boundary of the atmosphere. The aerosol is assumed to be gray, that is, τ_a does not depend on frequency in the range of the spectrum section being considered, 670-820 cm^{-1} . Also, only models of aerosol stratification were considered in which the entire aerosol was concentrated in a comparatively thin uniform layer between any two levels in the atmosphere. Such a model makes it possible to discover the relationship of the value of the aerosol effect to the altitude of the layer. /9

On the numerical value of $I_{\Delta V}$, random measurement error ϵ_V is "applied": $\bar{\epsilon}_V = 0$, $\sqrt{\epsilon_V^2} = 1 \text{ erg/cm}^2 \text{ cm}^{-1} \text{ steradian second}$, which amounts to 1-2% of the value of $I_{\Delta V}$, depending on V .

The inverse problem was solved according to measurements of the values of $I_{\Delta V}$ -- the profile of $T(\xi)$ was established calculating only absorption of carbon dioxide. The establishment was conducted by a method of statistical regulation [14] in two variations: 1) along the 670, 690, 702, 730 cm^{-1} channel and 2) along the entire sixth channel. Then, statistical probability of the temperature field from [16] was used (Berlin, summer). The method of statistical regulation frequently uses

probability information for solution, and probably is the most reliable as far as error in assigning the kernel of the integral equation in comparison with methods used with a smaller volume of probability information (see, for example, [17]). The profiles of $T(\xi)$ for which the calculations were made were selected in the form of expansions along the intrinsic vector of the correlation matrix of the temperature field $B_{TT}(\epsilon_1, \epsilon_j)$:

$$t(\xi) = T(\xi) - \bar{T}(\xi) = \sum_{i=1}^n c_i \varphi_i(\xi) \quad (4)$$

Here t is deviation of a given profile from the average, $\varphi_1(\xi) - 1$ is the intrinsic vector, C_1 is random numbers distributed according to normal law: $\bar{C}_1 = 0$, $\bar{C}_1^2 = \lambda_1$, λ_1 is the 1th intrinsic number of the matrix B_{TT} . Calculations were made for 12 realizations of $T(\xi)$, including the average profile of $t_0(\xi) \equiv 0$. Precision in establishment was determined according to two parameters, the mean quadratic error of establishment σ_{est} for all levels and error in establishment of temperature of the underlying surface Δt (1).

It is well known that the precision of the statistical regulation method with the same measurement errors is identical for different profiles of $t(\xi)$. It is the best for profiles close to the average and decreases as one gets farther from the mean profile. Therefore, the results of numerical experiments will be demonstrated for an example of establishing three different profiles including mean climate $\bar{T}(\xi)$. These profiles are established by a closed system with different precision. Δt (1) and σ_{est} for them for both variations of the solution (according to a closed system) are presented in table 1. If the mean climatic profile is established according to a closed system along the sixth channel with $\sigma_{est} = 0.3K$, then the profile of $t_2(\xi)$ is established under the same conditions with $\sigma_{est} = 2.76K$. Such an establishment error is

close to the upper limit of possible errors in establishing t (ξ) according to the closed system. One notes that the use of the 700 and 8[illegible] cm^{-1} channels, each of which is on the edge of a band, provides stability for establishing t (I) according to a closed system in all cases, inasmuch as in these channels, the atmosphere is very transparent and radiation in them is formed in the ground layer of the atmosphere.

Four cases of location of an attenuated layer are considered:

- a) below 800 mb;
- b) between 650 and 500 mb;
- c) between 100 and 300 mb;
- d) between 100 and 200 mb.

Figure 1 shows profiles of: T_0 , T_1 and T_2 and the position of the layer in all four cases.

The relationship of σ_{est} and $\Delta t(1)$ to τ_a in a case where the layer is between 400 and 300 mb for t_0 , t_1 and t_2 , is shown in figure [illegible] and 2b. In all cases $\Delta t(1)$ depends on τ_a much more strongly than σ_{est} . This attests to the fact that the aerosol effect occurs more strongly as a whole in the lower layers of the atmosphere. Using the 830 and 760 cm^{-1} channels provides a stable establishment of $\Delta t(1)$ in a case where the /11 kernel is precisely known and at the same time results in increasing the relationship to τ_a with the presence of additional absorption. It follows from figure 2 that the absolute magnitude of the aerosol effect hardly depends on the distance of the true profile from the mean. In cases of profiles of t_0 and t_1 , the course of σ_{est} , as a function of τ_a , is approximately uniform. The weak relationship of σ_{est} to τ_a in the case of τ_2 is explained by the fact that in this case, the temperature drop between the layer and the underlying surface is

somewhat smaller. With an increased altitude in the position of the layer (corresponding to a decrease in its temperature), its effect increases. Table 2 shows the value of τ_a necessary for a change in σ_{est} and Δt (1) of 1° Kelvin in each case. The tracings of σ in the graph for levels of the 1000-800 mb layer mean that even at $\tau_a = \infty$ (completely nontransparent layer) precision in establishing, on the average, changes less than at $1^\circ K$. At the 200-100 mb level for this, in order to worsen the establishment, it is adequate to have a layer with thickness 0.06-0.1. For discovering the ground temperature 0.02-0.03 thickness is adequate. One can assume the existence of layers with such thickness at the level of the tropopause. This can be [illegible] smoke for example.

The relationship of errors in establishment from the level of position of the layers is illustrated in figure 3 (a and b). The errors indicated of establishing the average profile for various altitudes of the layer with negligible optical thickness $\tau_a = 0.1$. Measurement errors in this case were zero so that the deviation indicated from the mean profile ("true") is explained only by the [illegible] layer. Here, errors in establishing the mean profile $t_0(\xi)$ was established according to [illegible] system with measurement error ϵ_v at 1-2%. It is apparent from the drawing that in this case, when the aerosol is concentrated in the low layers (curve 2), its effect on establishing $t(\xi)$ is not great.

For increasing the altitude of the layer, errors in establishing their entire levels increase. One finds /12 temperature in the lowest layers strongest of all. It is apparent that in this case, when the aerosol is concentrated close to the ground surface (between 1000 and 800 mb) additional attenuation does not occur. The layers at the level of the tropopause seek the temperature of the ground surface at almost $5^\circ K$. Moreover, when using the 700 and 820 cm^{-1} channels, errors in establishing increase somewhat.

From what has been presented above, one can make the following conclusions.

1. The actually existing aerosol formations can significantly decrease precision in establishing a temperature profile in the atmosphere. The high aerosol layers have a particularly strong effect.

2. The use of sections of the spectrum lying at the edge of the band in a murky atmosphere cause a decrease in precision in measuring the inverse problem.

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Table 1.

/15

Mean Quadratic Errors in Establishment and Errors in
Establishing Temperature of the Underlying Surface
(2 K) of the t_0 , t_1 and t_2 Profiles (Closed System
 $v = 1\text{ erg/cm}^2\text{.s.cm}^{-1}\text{.steradian}$).

	t_0		t_1		t_2	
	ϵ	$\Delta t(I)$	ϵ	$\Delta t(I)$	ϵ	$\Delta t(I)$
4 Channels	0.43	0.76	1.74	1.79	3.09	2.03
6 Channels	0.30	0.56	1.72	-0.23	2.76	0.81

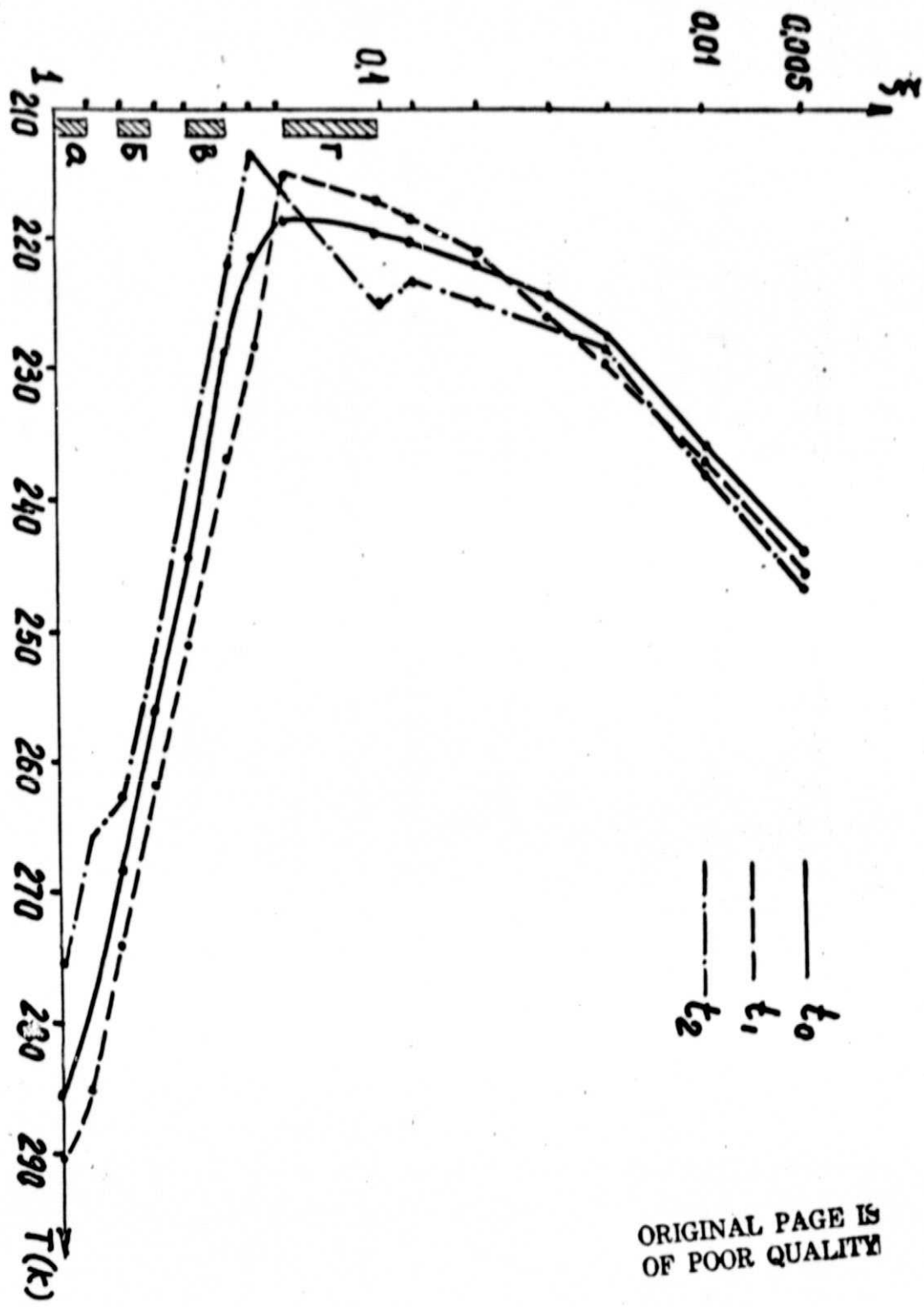
Table 2

Optical Thicknesses τ_a at which in each of the cases considered σ_{est} and $t(1)$ change on the infrared.

Position of the layer	t_0		t_0		t_1		t_1		t_2		t_2	
	6 channels	4 channels	6 channels	4 channels	6 channels	4 channels	6 channels	4 channels	6 channels	4 channels	6 channels	4 channels
	σ_{est}	$\Delta t(1)$	σ_{est}	$\Delta t(1)$	σ_{est}	$\Delta t(1)$	σ_{est}	$\Delta t(1)$	σ_{est}	$\Delta t(1)$	σ_{est}	$\Delta t(1)$
1000-300 mb	-	0.33	-	0.84	-	0.43	-	0.56	-	0.32	-	0
650-500 mb	0.19	0.04	0.27	0.06	2.20	0.08	0.30	0.08	0.48	0.03	0.40	0
400-300 mb	0.095	0.02	0.1	0.04	0.80	0.03	0.11	0.03	0.14	0.03	0.15	0
200-100 mb	0.055	0.02	0.07	0.02	0.055	0.02	0.03	0.02	0.13	0.03	0.14	0

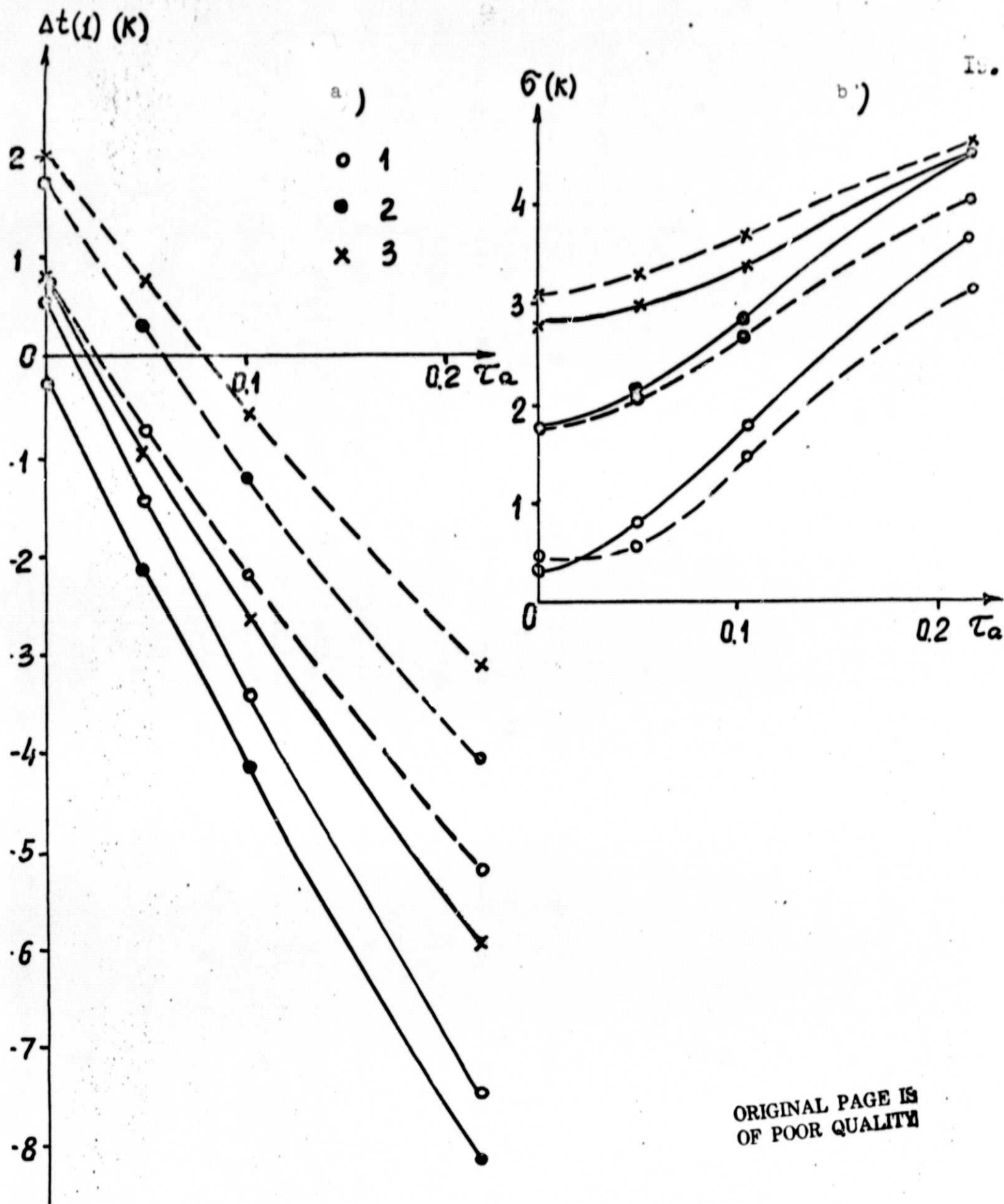
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1. T_0 , T_1 and T_2 profiles and the position of the aerosol layer in each case.
 - a) Layer between 800 and 1000 mb (1000 mb - is the level of Earth.
 - b) Layer between 650 and 500 mb.
 - c) Between 400 and 300 mg.
 - d) Between 200 and 100 mb.
2. Relationship $\Delta t(1)$ (2a) and σ_{est} (2b) when establishing the profile of t_0 (1), t_1 (2) and t_2 (3) according to the sixth channel (solid line) and according to the fourth channel (dashed line).
3. Establishing the average profile of t_0 (ξ) with the presence of a layer with $\tau_a = 0.1$ ($\epsilon_v \approx 0$) and according to the sixth channel (3a) and the third channel (3b).
 - 1 - closed system (measurement error 1%)
 - 2 - Layer below 800 mb.
 - 3 - Between 650 and 500 mb.
 - 4 - Between 400 and 300 mb.
 - 5 - Between 200 and 100 mb.



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Figure 1



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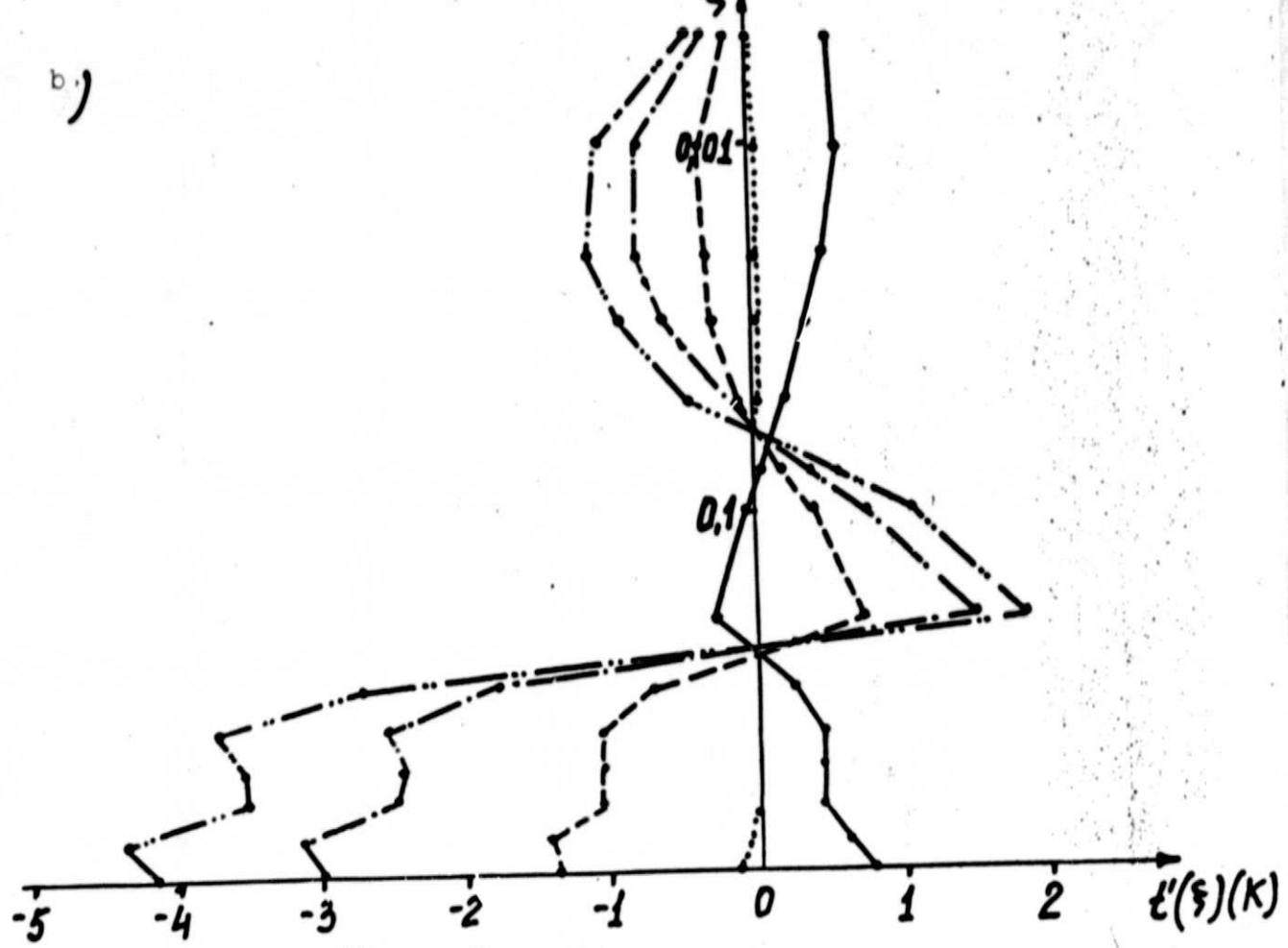
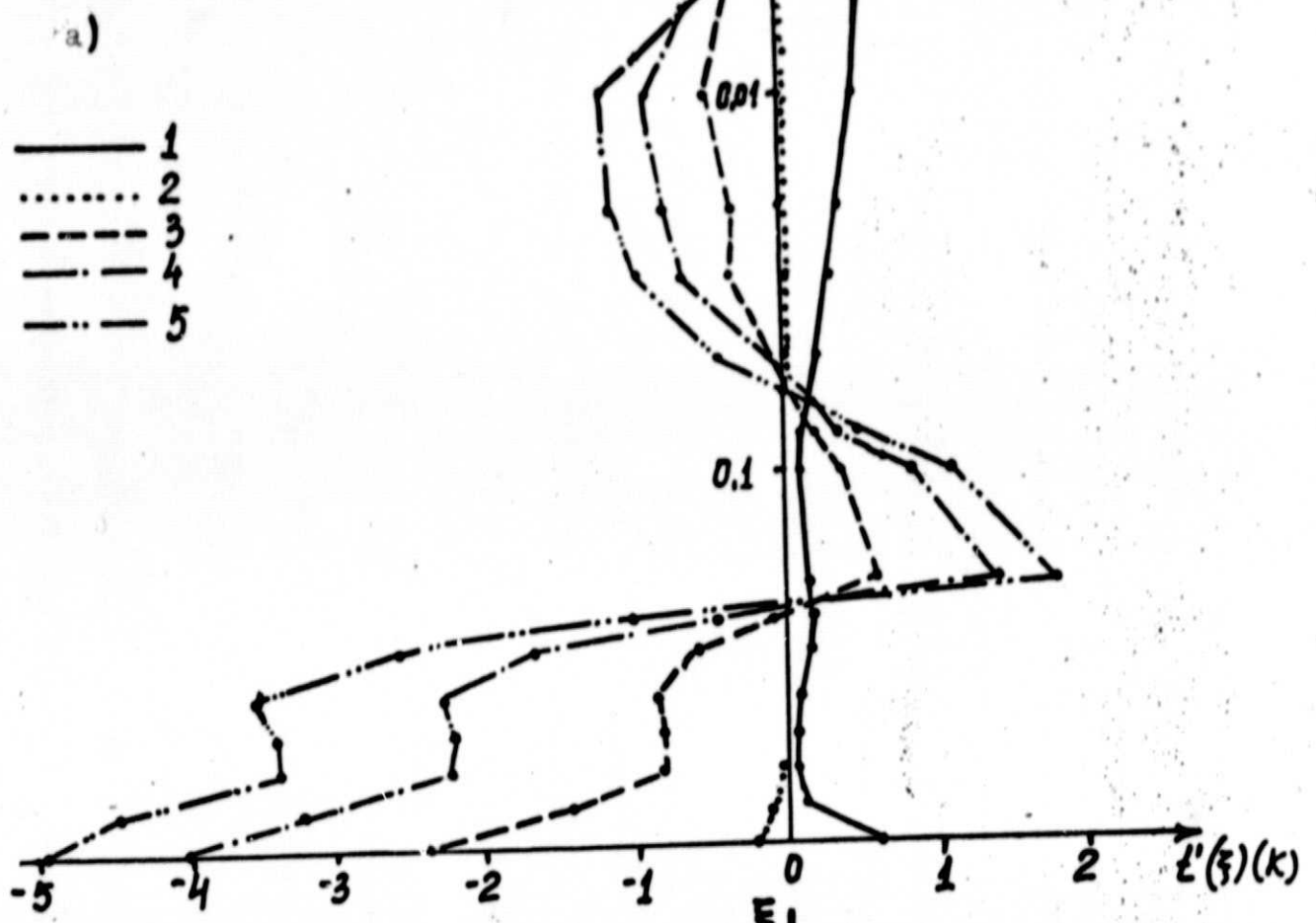


Figure 3

Transmission Measurements of Radiation in "Transparency Windows" of the Ground Layer of the Atmosphere.1. Introduction

Measurements of the coefficients of attenuation of radiation σ_λ of the ground layer of the atmosphere in 0.37 - 13 μm spectrum range were made in order to study the variations of transmission in "transparency windows" of the atmosphere and also for evaluating the contribution of water vapor and aerosols in the continual attenuation. The meteorological parameters of the layer were measured at the same time: Temperature t and partial pressure of water vapor e (or relative moisture r).

Measurements were made for various seasons of the year which made it possible to encompass a broad range of changes σ_λ , t , e and r and to establish the empirical relationship of σ_λ to meteorological parameters.

Examples of typical realizations of σ_λ , t , e , r and moisture content on the course are presented in table 1.

2. Measurement Methods

Measurements of the spectral transparency of the atmosphere in the 2-3 μm range were presented using an IK-20 dual-beam spectrometer with resolution along the 2-4 cm^{-1} spectrum. A glow bar was used as the radiation source. A system of mirrors provided passage of the beam on a course 1300 m long. An example of registering spectra of transmission is presented in figure 1.

A filter photometer graduated in absolute transmission values was used for measuring transmission in the 0.37-2.2 μm range on this same course. The interferential filters were made in eight sections of the spectrum with centers at 0.37, 0.42, 0.50, 0.63, 0.78, 1.03, 1.63, 2.23 μm (the half width of the filters was 0.01 μm for the first six filters and 0.04 μm for the last two). /22

Simultaneous measurements of transmission in the 2.3 μm "window" by two probes made it possible to make absolute coefficients of attenuation in the entire infrared range of the spectrum.

Measurement errors σ_λ do not exceed 0.03 km^{-1} .

3. Results of Statistical Processing

As a result of the statistical processing of σ_λ measurements, mean values of the attenuation coefficient were obtained $\bar{\sigma}(\lambda_k)$, mean quadratic deviations $d_\sigma(\lambda_k)$ (figure 2) and coefficients of correlation $R_{\sigma\sigma}(\lambda_k, \lambda_e)$ for the λ_k wavelength corresponding to the "transparency window" (table 2).

These statistical characteristics make it possible in the best way to determine the σ_λ "window" of the entire spectral range by measuring the coefficient of attenuation in one or several wavelengths.

$R_{cc}(\lambda_e, \lambda_e)$

Table 2. Standardized Correlation Matrix

λ , μm	0,37	0,50	1,03	1,6	2,2	3,16	3,7	8,33	8,63	8,96	9,33	10,2	10,53	10,9	11,6	12,7	13,05
0,37	I,00	0,93	0,85	0,78	0,79	0,69	0,65	0,65	0,68	0,69	0,68	0,65	0,68	0,67	0,71	0,70	0,67
0,50		I,00	0,94	0,85	0,78	0,79	0,68	0,64	0,63	0,67	0,66	0,63	0,67	0,67	0,70	0,69	0,68
1,03			I,00	0,92	0,84	0,84	0,74	0,70	0,67	0,70	0,69	0,69	0,74	0,74	0,74	0,73	0,73
1,6				I,00	0,92	0,88	0,80	0,79	0,76	0,79	0,77	0,77	0,81	0,81	0,81	0,82	0,82
2,2					I,00	0,94	0,84	0,86	0,84	0,86	0,84	0,86	0,88	0,87	0,87	0,87	0,88
3,16						I,00	0,88	0,88	0,86	0,88	0,87	0,88	0,88	0,87	0,89	0,88	0,89
3,7							I,00	0,85	0,84	0,86	0,86	0,88	0,87	0,86	0,83	0,82	0,82
8,33								I,00	0,94	0,95	0,94	0,94	0,93	0,90	0,96	0,92	0,92
8,63									I,00	0,95	0,95	0,93	0,90	0,88	0,92	0,90	0,87
8,96										I,00	0,98	0,95	0,90	0,88	0,95	0,92	0,90
9,33											I,00	0,96	0,92	0,90	0,95	0,93	0,91
10,2												I,00	0,96	0,94	0,96	0,94	0,93
10,53													I,00	0,97	0,95	0,94	0,94
10,9														I,00	0,95	0,92	0,93
11,6															I,00	0,96	0,96
12,7																I,00	0,97
13,05																	I,00

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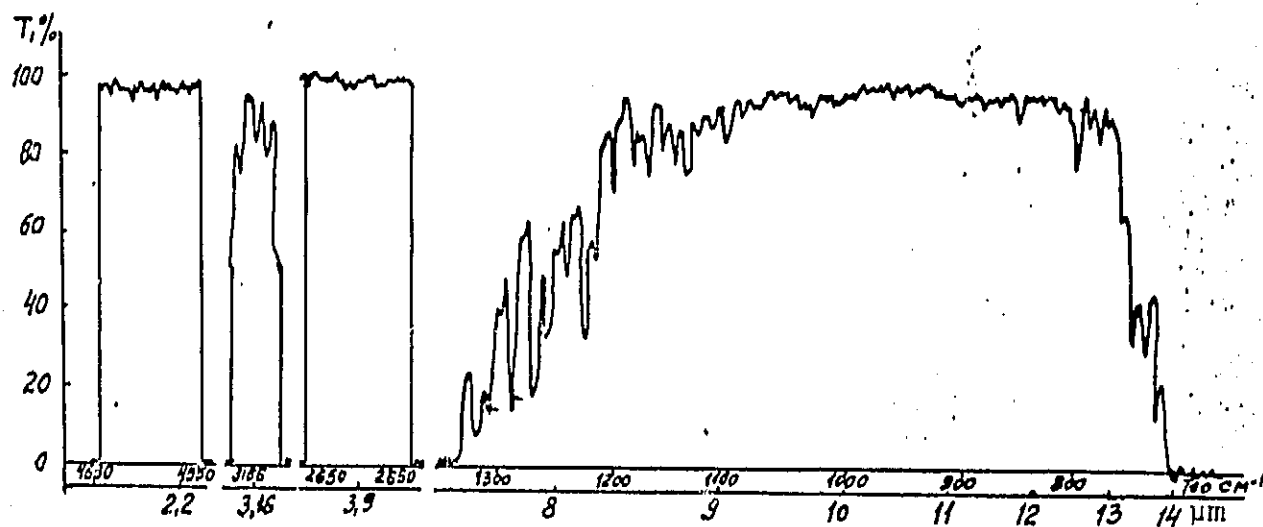


Figure 1. Recorded transmission of the atmosphere on a horizontal course. Recording was obtained on February 20, 1973, with a high transparency of the atmosphere. The value of the precipitated layer of water vapors amounts to $0.19 \text{ g/cm}^2 \text{ km}$.

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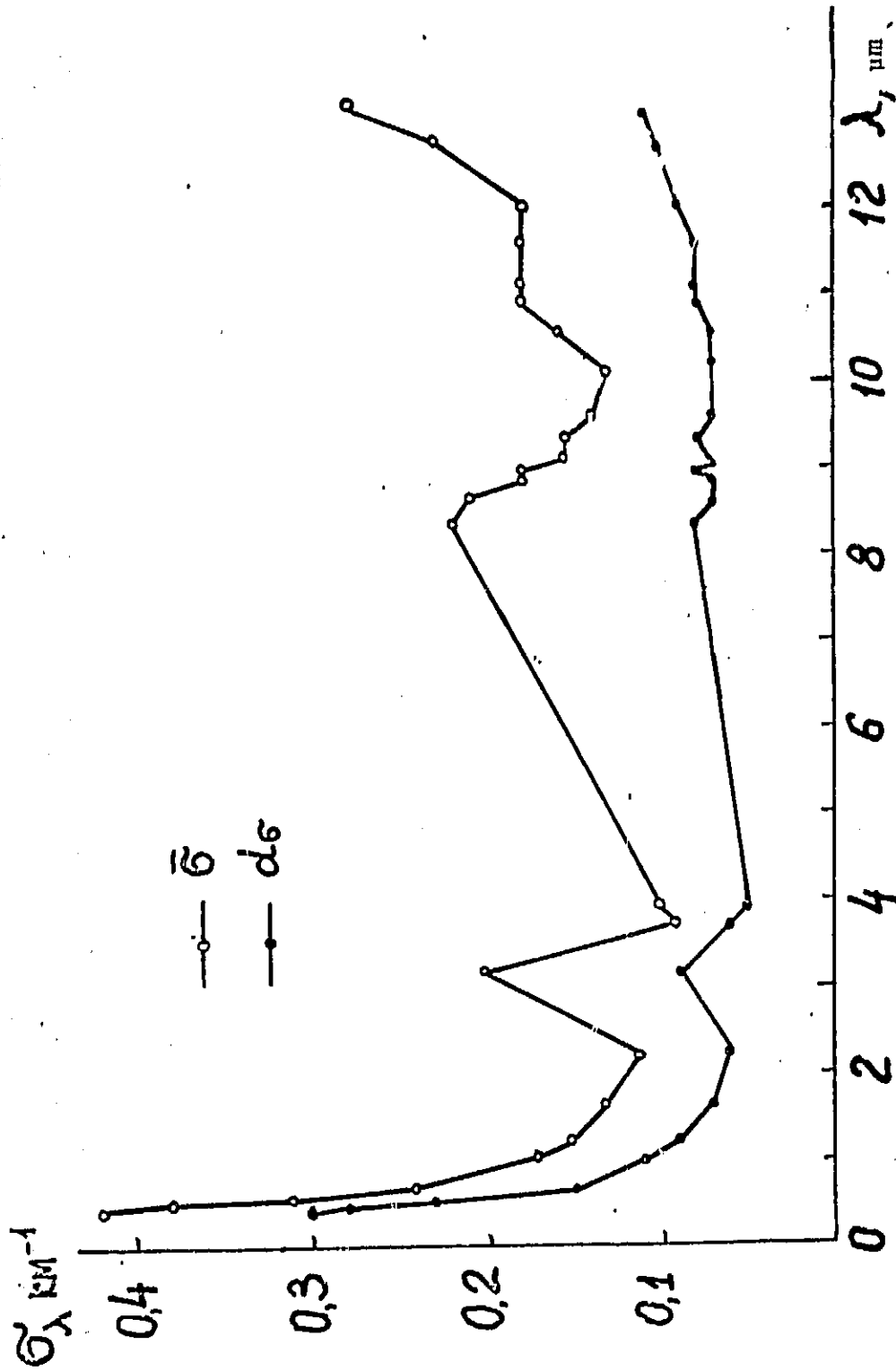


Figure 2. The spectral course of the mean coefficient of atmospheric attenuation σ_λ and their mean quadratic deviation $d_\sigma(\lambda_k)$.

Table 1. Examples of typical realizations of the spectral course of coefficients of radiation attenuation σ_{λ} (km^{-1}) in "transparency windows" with a ground layer of the atmosphere.

Date	Time	Wavelength, μm																							
		1.6	2.2	3.6	3.7	3.9	6.3	6.4	6.5	6.6	6.7	6.8	6.9	7.0	7.1	7.2	7.3	7.4	7.5	7.6	7.7	7.8	7.9	8.0	8.1
15.03.73	10 ^h .00 ^m	0.37	0.42	0.50	0.63	0.78	1.03	1.2	1.6	2.2	3.6	3.7	3.9	6.3	6.4	6.5	6.6	6.7	6.8	6.9	7.0	7.1	7.2	7.3	7.4
12.03.73	10 ^h .00 ^m	0.08	0.06	0.04	0.03	0.05	0.04	0.02	0.03	0.05	0.01	0.03	0.02	0.06	0.06	0.06	0.06	0.06	0.06	0.06	0.06	0.06	0.06	0.06	0.06
31.10.72	19 ^h .30 ^m	0.31	0.28	0.23	0.22	0.20	0.16	0.13	0.14	0.14	0.30	0.19	0.16	0.14	0.16	0.20	0.17	0.17	0.14	0.13	0.13	0.13	0.13	0.13	0.13
25.10.72	22 ^h .30 ^m	0.31	0.28	0.24	0.19	0.13	0.10	0.07	0.06	0.05	0.14	0.05	0.06	0.12	0.14	0.12	0.12	0.11	0.06	0.06	0.06	0.06	0.06	0.06	0.06
16.10.72	20 ^h .30 ^m	0.20	0.17	0.13	0.12	0.11	0.11	0.11	0.09	0.07	0.13	0.05	0.06	0.08	0.19	0.23	0.19	0.16	0.16	0.16	0.16	0.16	0.16	0.16	0.16
27.09.72	21 ^h .30 ^m	0.31	0.28	0.22	0.22	0.19	0.16	0.14	0.12	0.12	0.22	0.11	0.11	0.12	0.26	0.27	0.23	0.20	0.19	0.19	0.19	0.18	0.16	0.17	0.19
13.09.72	20 ^h .30 ^m	0.80	0.74	0.59	0.50	0.44	0.32	0.27	0.22	0.19	0.30	0.19	0.18	0.19	0.35	0.37	0.32	0.28	0.27	0.26	0.25	0.24	0.23	0.22	0.27
12.10.71	23 ^h .30 ^m	0.58	0.51	0.45	0.32	0.30	0.26	0.27	0.18	0.18	0.29	0.10	0.10	0.11	0.29	0.34	0.28	0.26	0.26	0.25	0.22	0.23	0.19	0.16	0.16
25.09.72	23 ^h .30 ^m	0.62	0.54	0.36	0.34	0.20	0.18	0.14	0.10	0.10	0.19	0.10	0.09	0.10	0.24	0.27	0.23	0.19	0.16	0.15	0.16	0.16	0.16	0.16	0.16
25.09.72	22 ^h .00 ^m	0.42	0.39	0.26	0.20	0.16	0.16	0.14	0.10	0.10	0.19	0.09	0.09	0.10	0.20	0.23	0.19	0.16	0.16	0.16	0.16	0.16	0.16	0.16	0.16

Values of Meteorological Parameters

Date	Time	Wavelength, μm																							
		1.6	2.2	3.6	3.7	3.9	6.3	6.4	6.5	6.6	6.7	6.8	6.9	7.0	7.1	7.2	7.3	7.4	7.5	7.6	7.7	7.8	7.9	8.0	8.1
15.03.73	10 ^h .00 ^m	-2.4	1.8	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9
12.03.73	10 ^h .00 ^m	-5.0	3.8	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9
31.10.72	19 ^h .30 ^m	-0.4	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9
25.10.72	22 ^h .30 ^m	4.0	6.0	6.0	6.0	6.0	6.0	6.0	6.0	6.0	6.0	6.0	6.0	6.0	6.0	6.0	6.0	6.0	6.0	6.0	6.0	6.0	6.0	6.0	6.0
16.10.72	20 ^h .30 ^m	5.2	6.3	6.3	6.3	6.3	6.3	6.3	6.3	6.3	6.3	6.3	6.3	6.3	6.3	6.3	6.3	6.3	6.3	6.3	6.3	6.3	6.3	6.3	6.3
27.09.72	21 ^h .30 ^m	6.8	9.1	9.1	9.1	9.1	9.1	9.1	9.1	9.1	9.1	9.1	9.1	9.1	9.1	9.1	9.1	9.1	9.1	9.1	9.1	9.1	9.1	9.1	9.1
13.09.72	20 ^h .30 ^m	15.0	11.1	11.1	11.1	11.1	11.1	11.1	11.1	11.1	11.1	11.1	11.1	11.1	11.1	11.1	11.1	11.1	11.1	11.1	11.1	11.1	11.1	11.1	11.1
12.10.71	23 ^h .30 ^m	12.6	13.2	13.2	13.2	13.2	13.2	13.2	13.2	13.2	13.2	13.2	13.2	13.2	13.2	13.2	13.2	13.2	13.2	13.2	13.2	13.2	13.2	13.2	13.2
25.09.72	23 ^h .30 ^m	13.0	14.1	14.1	14.1	14.1	14.1	14.1	14.1	14.1	14.1	14.1	14.1	14.1	14.1	14.1	14.1	14.1	14.1	14.1	14.1	14.1	14.1	14.1	14.1
25.09.72	22 ^h .00 ^m	14.0	15.6	15.6	15.6	15.6	15.6	15.6	15.6	15.6	15.6	15.6	15.6	15.6	15.6	15.6	15.6	15.6	15.6	15.6	15.6	15.6	15.6	15.6	15.6

Notation: t - temperature, e - partial pressure of water vapor, r - relative humidity, W - precipitated layer of water vapors.

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16. Abstract This article contains a survey of literature on errors in establishing temperature of the underlying surface and the vertical profile of temperature in the 15 μ m CO ₂ band when aerosol attenuation was not taken into account. Experimental data show the large contribution of the aerosol to total radiation attenuation. Evaluations are made of the relationship of the magnitude of establishment error to location and optical thickness of the aerosol layer.			
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